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Secondary Frontal Waves

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SECONDARY FRONTAL WAVES

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1 Introduction

Although interest in frontal waves, as instabilities of midlatitude frontal zones, goes back to the times of the Bergen school, with subsequent studies in later years, significant advances have recently been made, since the late 1980's. The idea of secondary frontal waves stems from a duality in the understanding of frontal instability, which again dates to the times of the Bergen school. In midlatitudes, cyclonic systems develop as baroclinic instabilities of the meridional temperature gradient, or 'polar front', as typified by the Charney and Eady models of baroclinic instability. Within these baroclinic waves, the action of the shearing of the basic state is to form locally intense fronts which will reach small lengthscales in finite time (as shown by Hoskins and Bretherton 1972). Recently it has been noted that the intense fronts associated with such developed baroclinic waves may themselves undergo instability, forming secondary cyclones, for example, on the trailing cold front. These secondary systems occur on the mesoscale, can have extremely large growth rates, developing over one or two days, and are notoriously hard to forecast. The recent studies of frontal waves focus on these secondary instabilities, although it could be argued that the difference between the secondary systems and their parent instabilities is only one of scale.

Detailed observations of explosive cyclogenesis have recently taken place in the ERICA project of 1988/1989, studying rapidly intensifying cyclones over the West Atlantic, close to the East coast of the USA. Although these systems develop rapidly on the baroclinicity between the Atlantic and the North American continent, with small horizontal scales, it may be argued that they represent 'primary' cyclogenesis, as they occur at the entrance to the Atlantic storm track. In any case, the planetary scale flow feeding these systems, with westerlies from the continent flowing over the warm Gulf Stream water, provides different conditions for cyclone growth than the mature fronts which approach Northern Europe. Other rapidly developing mesoscale cyclones include 'polar lows', which develop as cold polar air flows over high latitude oceans and are often associated with ice-boundaries. Again, these phenomena are here regarded to be different from the secondary frontal cyclones which commonly affect the UK and France.

The issues at stake are the energy source for the development (whether baroclinic, as the primary synoptic frontal cyclones are known to be, or barotropic as some studies have found),

how intense or 'explosive' growth rates occur, the role of moist processes (in conditioning the primary front and in influencing the secondary system) and the structure of the mature frontal wave.

2 Recent observational studies

Observations of secondary frontal waves have been hampered by the same factors which make them difficult to forecast; their small scales and their rapid growth rates, although retrospective studies of significant events such as the 'October Storm' of 1987 (Hoskins and Berrisford 1988, Shutts 1990) have proved revealing. Despite such difficulties, recent study of ECMWF analyses by Ayrault et al. (1995) (ALJL) and an observational program carried out by the JCMM, FRONTS 92, have greatly improved our understanding of the structure of these phenomena.

ALJL used ECMWF 1984-1994 cold season analyses for the North Atlantic region, at resolution of T106 for the early part of this period and T213 after September 1991. Thus, the scale of resolution before 1991 was around $400km$; statistical analysis of the results suggests that there is no qualitative change in the behaviour of analysed frontal waves due to the change in resolution. This study first identified four weather regimes for the large scale flow, within which higher frequency disturbances would be located by an objective method: these regimes concurred, in the main part, with regimes diagnosed elsewhere in the analysis of primary baroclinic waves. ALJL make the point that the high frequency signal used to locate frontal waves may embrace many phenomena such as polar lows, Mediterranean cyclones and 'ordinary' frontal waves, and this regime-dependent analysis is intended to differentiate between the qualitatively different phenomena. Within the individual weather regimes, the high frequency signal is maximised at the exit regions of the storm tracks, confirming that the high frequencies represent systems which grow on decaying baroclinic waves. Also, this observation indicates that the rapidly-growing frontal waves of the East Atlantic are of a different nature to the systems which have been studied in the Western Atlantic, off the East coast of the USA (to be discussed later). ALJL's study focusses on the 'Zonal' weather regime, in which maximum high frequency activity occurs in the region of France and the UK. In this regime, the two principal groups of disturbances are labelled 'type 1' and 'type 2' and have distinct characteristics; although there

are some problems inherent in the averaging involved with this work, these types of secondary cyclone offer a generality to the study of frontal waves: they are described in detail.

Both types have a characteristic scale of $1,000km$. Type 1 cyclones exhibit a prominant cold front, which reinforces the parent frontal zone. In contrast, type 2 cyclones have a stronger warm front. Both types show a relatively weak pressure signal of order $3mb$ and a temperature wave of amplitude $2.5K$: ALJL suggest that the clustering technique may be responsible for ameliorating the amplitude of the resulting structures. Both types exhibit clear signatures in the relative humidity field: in type 1 cyclones this feature is evident well before the reference time while in type 2 it is weak at $-6h$ and intensifies rapidly as it spreads over the warm front. The warm anomaly of type 1 is colocated with the relative humidity maximum while for type 2 the relative humidity at $700mb$ is shifted ahead of the $850mb$ warm maximum (as in Joly and Thorpe 1990).

Potential vorticity analysis in conjunction with these studies shows related high frequency behaviour and it is shown that despite having weak pressure signals, these systems do have an upper level ($300mb$) component on a similar scale. In type 1 cases the upper level anomaly can be traced back $18h$ but type 2 cyclones have little signal at $-6h$, in low level θ_e or $300mb$ pv.

In terms of the large scale flow, type 1 cyclones appear to develop in an environmental deformation field which is weakly frontogenetic. Type 1 form on a well-defined cold front which is reinforced by the wave's cold front. In the case of type 2, the background flow is frontolytic and the background baroclinicity is weaker than in type 1. This observation pertains to the discussions of Bishop (1993a,b) and Bishop and Thorpe (1994a,b) on the role of deformation frontogenesis in frontal wave growth, discussed subsequently.

Energy calculations show that both types of cyclone have dominant baroclinic sources at lower levels. Type 1 also shows a positive barotropic contribution but type 2 cases lose energy to this term. Type 1 cyclones also gain energy from upper level baroclinicity.

In summary, the type 1 cyclone is a mixed baroclinic/barotropic wave with upper level interaction. Type 2 is predominantly baroclinic with an essentially low level source. ALJL suggest that the type 1 case may be described by familiar concepts of baroclinic development but that type 2 cyclones represent systems which are largely dominated by the background frontolytic deformation field, a scenario which requires further theoretical investigation.

Many features of the IOP3 case study of the FRONTS 92 project (Hewson, 1993) concur with the analysed 'type 2' disturbance of ALJL. FRONTS 92 was an observational programme which made use of the MRF C130 aircraft's dropsondes and airborne radar and ground based observing systems to observe secondary frontal waves in early 1992. IOP3 was the most successfully observed case, and developed rapidly, southwest of the UK, growing by $12mb$ in the 12 hours of intense observation by the aircraft. The horizontal scale of the system was around $500km$. As in ALJL's type 2 case, the synoptic situation was one of diffluent, frontolytic flow downstream of a col over the West Atlantic. However, this wave was one of a succession of disturbances to a trailing cold front over the Atlantic and could be traced back for over 2 days, when genesis appeared to occur, in a region of very intense SST gradients (around $4K(100km)^{-1}$; even more intense SST gradients may occur in this region, as seen in Reed *et al.* 1993a). There is evidence from UKMO LAM output that the interaction with an upper level trough caused the growth of the wave on 27/28 April as it approached the UK.

Pronounced features of this wave, as observed in the dropsonde data, include a pronounced warm front (as in ALJL's type 2) and at least two cold fronts, as discussed by Browning *et al.* (1995); large vertical wind shears outside the warm sector and a temperature range of order $10K$ through the wave. As with ALJL's type 2 cyclone, the relative humidity anomaly of IOP3 seemed to lie ahead of the warm sector; this was concurrent with an elevated ($\sim 700mb$) cold θ_w front overrunning the surface warm front (described by Browning *et al.* 1995). A rainband, which persisted for at least $15h$, lay parallel to the warm front and ahead of it, and seemed to be riding up the warm frontal surface.

The warm frontal flow as derived from the dropsonde data (Hewson, 1993) is sketched in figure 1. The rainband was associated with a 'kink' in the frontal surface, where the equivalent potential temperature was constant with height; similar kinks were observed by Locatelli *et al.* (1994) for shorter-lived cold frontal rainbands. Locatelli *et al.* (1994) ascribe the rainbands as being forced by a trapped gravity wave between their cold frontal surface and the ground, basing their argument on the nonhydrostatic, nongeostrophic simulation of dry deformation frontogenesis performed by Gall *et al.* (1988). Gall *et al.* (1988) associated stationary (relative to the front) patterns in the cross-frontal flow and frontal topography with such trapped gravity waves and showed a close link with theoretical scales expected for such waves. The effect of

latent heat sources and sinks on these waves is not explored but might be expected to reinforce their structure: in particular, evaporative cooling of precipitation may interact to give a positive feedback on the intense and narrow downdraughts found beneath the frontal kink. However, other self-consistent arguments could be invoked to explain the coupling between the frontal kink and the rainband: in IOP3 the rainband persisted for at least $15h$ and lay above a maximum in the along-front wind ('cold conveyor belt') associated with warm advection; it is likely that balanced processes had a bearing on the propagation. Parker and Thorpe (1995) show how a precipitation band in a semigeostrophic model tends to lie above the warm low level jet to the flank of the low level pv maximum in a frontal system. They describe a propagation mechanism for such a feature: a drawback in using the trapped gravity wave model to describe the IOP3 warm frontal rainband is that in contrast with the observations of Locatelli *et al.* (1994), the rainband propagates relative to the front. In Parker and Thorpe's (1995) model, the frontal kink would be seen as an effect of the diabatic heating and cooling through which θ_e becomes constant with height. A feedback on the jet at low levels may then occur through the local intensification of the horizontal temperature gradient, giving a greater vertical shear above the jet through thermal wind balance.

The relevance of these two physical models for the warm frontal rainband of IOP3 may lie in the question of whether the warm front is better described as a 'shear front' or a 'confluence front'. Since it is reasonable to hope that IOP3 is one of a generic type of frontal instability (such as ALJL's type 2), the environment of this warm frontal rainband may have a general occurrence—the interaction between this kink on the frontal zone and the rainband deserves further attention.

Browning *et al.* (1995) studied the multiple cold-fronts of IOP3 in greater detail. They describe the intrusion of dry air from behind the system, overrunning warmer, moister low level air of the warm sector. Browning and Golding (1995) show how a dry intrusion in a vigorous cyclone over the UK was involved in the generation of a tornadic squall line, and indicate that this dry air of high pv originated in the stratosphere rather than being formed at low levels by latent heat release. Again, aspects of these observations need greater theoretical understanding: the origin of the dry intrusions and the resulting banded structure which forms as they overrun the warm sector are not yet fully explained.

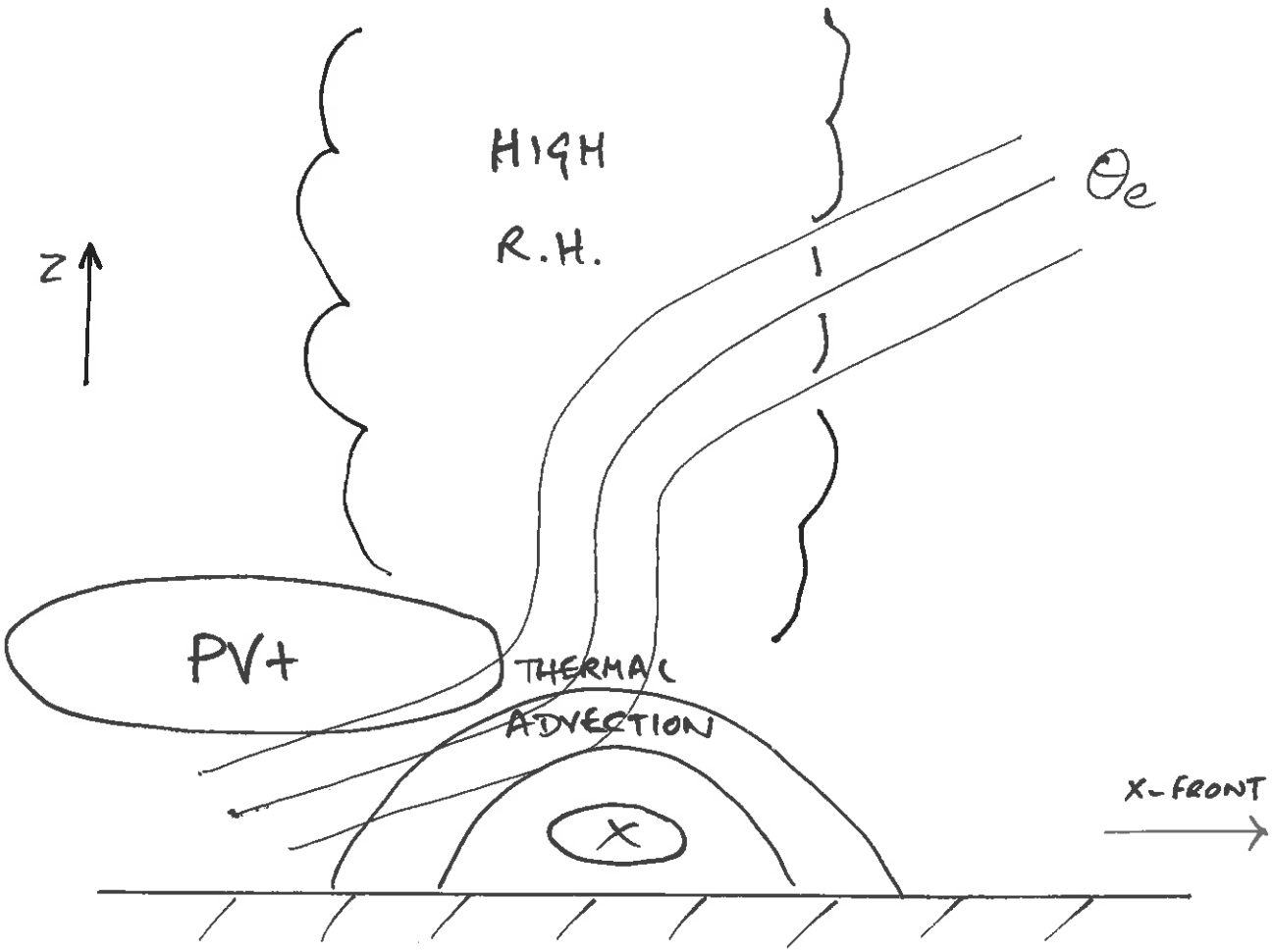

Figure 1: A sketch of the flow in the vicinity of the warm front of IOP3 of the FRONTS 92 project, as deduced from the dropsonde data.

Neiman, Shapiro and Fedor (1993) describe remarkably detailed observations of the mesoscale structure of an explosively developing cyclone ($60mb$ in $24h$) off the East coast of the USA. Their analysis shows frontal widths of $2km$ and frontal vorticity of up to $20 \times 10^{-4}s^{-1}$ at the convectively active fronts, and the authors compare these features with thunderstorm outflows (or 'cold pools') which are often described as gravity currents. These fronts exhibit shallow (below $750mb$) smaller scale vortices down to the $10km$ horizontal scale, which are clearly recognisable in a signature of deep convection. Although the context of these systems, occurring where cold continental air meets the warm boundary layer air over the Gulf Stream, is different than that of the cyclones of interest to European forecasters, it does seem that secondary frontal cyclones can exist down to the smallest resolvable scales.

3 Theoretical models

(a) The front

It is inevitable that a prerequisite of understanding frontal waves is an understanding of the front itself. Early dynamical descriptions of a front were based on the idea of 'airmasses', but observations do indicate that a front is in fact a transition zone rather than a discontinuity, and may propagate at a speed somewhat different from the that of the local wind. The semi-geostrophic view (Hoskins and Bretherton 1972) is able to describe these aspects of fronts with some success (Lagouvardos *et al.* 1993), provided diabatic processes are taken into account, and this report concentrates mainly on the discussion of instabilities to such non-discontinuous fronts.

The gravity current model of a front is often useful at smaller scales and has been proposed as a mechanism of instability or 'frontal fracture' (Hobbs and Persson 1982). However, the scales of relevance of gravity current dynamics, of order the depth ($1-2km$ in the atmosphere) and the importance of rotational effects means that although gravity currents may be useful for understanding the kind of small scale instabilities observed by Neiman, Shapiro and Fedor (1993) it is unlikely that this model will be useful for explanation of the larger frontal waves discussed here (scales of $500km$, lifetimes of days).

(b) Dry instabilities

Although it is now widely realised that normal mode analysis can fail to represent even the small amplitude fastest-growing structures, calculation of normal modes, as time-independent structures, can still be a revealing exercise. Schär and Davies, 1990, (SD) and Joly and Thorpe (1990) both presented work which developed normal mode instabilities of frontal regions.

For dry normal mode instability of balanced systems, the Charney-Stern theorem states that it is a necessary condition that the basic state pv gradient and the equivalent pv gradients which represent the boundary potential temperature gradients (Bretherton 1966) have regions of opposite sign. Schär and Davis (1990) and Joly and Thorpe (1990) introduce such pv gradient sign-changes into their models by assuming, respectively, a warm band ahead of the cold front and a strip of low level positive pv anomaly due to diabatic heating. It can be argued that, from the Bretherton analogy, these models have dynamical equivalence. At this stage it

should perhaps be observed that in flows where moist processes are significant, the Charney-Stern theorem need not be relevant. Snyder and Lindzen (1991) showed how the pv anomalies induced by diabatic heating in an infinite fluid (not containing pv gradient reversals) could act in a similar manner to anomalies induced by advection in a dry flow. Joly and Thorpe (1991) illustrate in another artificial model, with three dimensional instabilities to a front with no pv anomaly but retaining diabatic heating terms, how the Charney-Stern theorem may fail.

SD show how their warm band undergoes a mixed barotropic/baroclinic instability. A narrower warm band produces faster growth of the instability at a shorter wavelength. Similarly, a more pronounced temperature maximum produces faster growth. The instability is restricted to low levels. In the nonlinear regime, initialising with the normal mode, saturation occurs, leading to reduction in the growth rate from exponential. The warm band breaks up into isolated warm anomalies and the cold front develops a cold front/warm front structure. The doubling time of the waves is around 1.5 days.

Joly and Thorpe (1990) also used a dry semigeostrophic model to demonstrate the instability of a positive low level pv anomaly. They showed how the available energy for the instability depends on the shape of the anomaly, with a narrower pv strip having more available kinetic energy. As in SD, a narrow anomaly produces shorter, faster-growing waves, dominated in this case by the barotropic energy conversion. A deeper anomaly likewise produces faster growth and it would be of interest to see how the stability of a strip of pv, conserved in cross-section integrated intensity, varies if it is deformed. Joly and Thorpe (1990) also used a moist-modified Eady wave, as computed in Emanuel, Fantini and Thorpe (1987), as a steady basic state for the dry semigeostrophic instability of the low level pv. It is interesting to note that the modes found by Joly and Thorpe (1990) show vertical velocity and potential temperature perturbations to be out of phase at low levels, as seems to be the case in ALJL's type 2 (but not type 1) cyclone and IOP3 of FRONTS 92.

Subsequent work by Malardel *et al.* (1993) extended Joly and Thorpe (1990) by running nonlinear simulations in a primitive equation model. As in SD they found saturation of the modes. In fact, they noted that the growth is restricted so that the instability is insufficient to produce intense pressure changes in the developing lows, contrary to surface observations of many such systems. In terms of energetics, the barotropic conversion is initially dominant but

subsequently it decreases and the baroclinic conversion dominates; at this stage the barotropic conversion becomes negative and is a sink of energy and the cause of equilibration. Only the waves which go through this two-stage (barotropic then baroclinic) process deepen significantly. In order for this to occur, a suitable initial perturbation is required, deep enough to interact with the ambient baroclinicity, and sufficient ambient baroclinicity must be present. Malardel *et al.* (1993) also mention the inclusion of moist processes in similar studies and state that this gives little change from dry cases, although they give few details of these moist studies. In contrast, Ferris (1989) describes how frontal instability in a 3D mesoscale model can be dominated by the convective latent heating, and both Hoskins and Berrisford (1988) and Shutts (1990) suggest that latent heating was a critical factor in the intense development of the 'October Storm' of 1987.

Bishop (1993) explored analytical, non-modal solutions to the development of linear instabilities on a mixed, time dependent basic state of baroclinicity undergoing deformation frontogenesis. These methods were used by Bishop and Thorpe (1994a,b) (BTa,b) to investigate the influence of deformation frontogenesis on the growth of frontal waves. BTa showed how frontogenetic deformation acting on the low level vorticity strip at a front favours the growth of instabilities which are initially of short wavelength: the deformation field acts to increase the wavelength of the instability while increasing the potential growth rate by making the frontal vorticity strip narrower. However, BTb describe how a sufficiently large deformation will suppress nonlinear wave development. Bishop (1993) comments on the fact that *frontolytic* deformation can increase the growth rate of instabilities (for instance rendering initially neutral modes unstable) and the processes involved in the nonlinear regime of a frontolytic flow are certainly pertinent to the understanding of ALJL's type 2 cyclones, which seem to develop in such a large scale environment.

Bishop's work highlights the way in which the structure of a growing wave may change with time as it develops; it is now well established in the theory of cyclogenesis that 'non-modal' linear disturbances which change their form as they grow may experience much more rapid development than normal mode structures (Farrell 1984). Joly (1995) explored these ideas in the context of secondary frontal waves and showed how transient growth rates could be three times greater than those of the normal modes.

Thorncroft and Hoskins (1990) describe dry baroclinic wave lifecycle simulations on the sphere in which cyclogenesis was observed to occur on the trailing cold front as a result of the interaction with upper level cyclonic vorticity. This kind of finite amplitude interaction may be the way in which the instability *mechanisms* of normal mode solutions act transiently to give explosive growth, or a manner in which slowly-developing shallow waves may be reinforced. The wave observed during IOP3 of FRONTS 92 was one of a series of waves on a long cold front but the only one to acheive deep growth, and there is evidence that this was involved with an upper level trough. Such finite amplitude interactions may also be pertinent to the description of ALJL's type 1 cyclone, in which precursors to the development were evident at upper and lower levels for some time before the period of growth. Finite amplitude studies have a further advantage in that they can show the Lagrangian aspects of the flow, as small amplitude modes may fail to do. This aspect is important in studying the origin of the elements of an instability (for example the warm conveyor belt or a stratospheric intrusion). However, as for non-modal studies, it can be argued that finite amplitude models lack the generality of modal studies and, while providing a more precise description of an instability, may fail to illuminate its dynamical processes.

Although upper and lower level interactions are a cornerstone in the understanding of baroclinic instabilities, the short wavelengths of many observed frontal waves mean that the Rossby scale heights of disturbances will be small. Thinking in terms of canonical Eady wave dynamics, short waves may not grow from baroclinic instability. However, other canonical forms exist, such as the Charney model, where there is no rigid lid but a gradient of pv, Green's (1960) system which has a pv gradient and rigid lid (allowing short wave instability) or Snyder and Lindzen's (1991) model where an equivalent pv gradient is provided by diabatic sources; in addition, Joly and Thorpe (1990) show how a pv strip which is deep and narrow undergoes a principally *barotropic* instability. Given the range of baroclinic, diabatic and barotropic instabilities possible, it is perhaps unwise to focus attention solely on upper/lower level interactions as the source of rapid growth on the smaller scales, even from the perspective of transient development.

In summary, frontal instability has been seen as a growing disturbance on a pv strip (possibly enhanced by transient interactions with upper levels). The general energetics of the disturbance

depend on the cross-sectional aspect ratio of the pv strip (Joly and Thorpe 1990) and the wavelength and structure of fastest-growing normal modes depends on the width (narrower strip gives shorter, faster-growing waves). Deformation frontogenesis is seen to modify the development, as it deforms the developing disturbance and alters the width and stability of the basic front; the role of frontolytic flow has perhaps not been sufficiently explored. Moisture in the instability has been seen as the source for the basic pv strip, through the latent heating at the basic front, and many analyses have been of a dry disturbance to this pv distribution. The structure of such dry disturbances has been documented, and in the broadest sense these models concord with the limited observations of secondary frontal waves. The next section will describe what is understood about the role of moisture in the disturbances.

4 Moist instability

Although Malardel *et al.* (1993) state briefly that the inclusion of moist dynamics in the instability has little effect, Hoskins and Berrisford (1988), Ferris (1989) and Shutts (1990) would seen to contradict this, and there are some outstanding aspects of the moist dynamics which remain to be explored. Certainly, satellite images of frontal waves show that moist convection is taking place (Moore and Peltier, 1987). Polar lows (shallow intense cyclones that develop over high latitude oceans) are thought to have analogous dynamics to tropical cyclones, dominated by latent heating (e.g. Nordeng and Rasmussen 1992) and though these are not specifically the same phenomena as the frontal waves of interest here, it is likely that they share some features, or that there may be a spectrum of such cyclones, over which latent heating plays a differing role.

It is known that moist processes can significantly modify baroclinic instability (e.g. Emanuel *et al.* (1987), Craig and Cho (1988), Snyder and Lindzen (1991), Parker and Thorpe (1995)), especially when included in an 'unconditional' or 'moist up, dry down' manner, increasing growth rates and narrowing updraught scales. Since the growth of the dry modes of Malardel *et al* (1993) saturates and large pressure perturbations are not obtained, it would be reasonable to expect that the same moist parametrizations may similarly influence this instability.

In qualitatively assessing the role of moisture in a developing frontal wave it is important to

have an understanding of the velocity and thermodynamic fields associated with a perturbation to the basic pv strip, since this determines, interactively, the diabatic forcing of the instability. This interaction between the flow and the associated moist processes is a subject of much current discussion. Emanuel *et al.* (1994) have argued convincingly that in the context of large scale tropical disturbances it is misleading to link low-level moisture convergence with diabatic heating aloft. Instead, they review the theory that a convecting atmosphere is in a state of quasi-equilibrium in which the large-scale averaged convective inhibition is negligible and the adjustment by the convective elements is rapid compared to the timescale of the large scale forcing, so that the tropospheric temperature field on this scale is determined by the subcloud (moist) entropy distribution. At the other extreme—at the cloud scale—it is clearly demonstrated from routine observations that convective inhibition is common and in such situations convection must be 'triggered' by finite amplitude vertical displacements. At the mesoscale, as ever, the picture is much less clear. At midlatitude fronts there is evidence (e.g. Hobbs and Persson, 1982) that, as at squall lines, the finite amplitude upward displacement at the low level front acts to 'trigger' a band of intense convection. This is further justified by the diagnostic analyses of FRONTS 87 data performed by Lagouvardos *et al.* (1993) who show that a convective parametrization based on such assumptions is needed for a good representation of the cross-frontal flow. Contrary arguments by Thorpe and Emanuel (1985) would seem to indicate that the ascent region in the vicinity of a midlatitude front exists in a state close to moist slantwise neutrality. This controversy is not resolved and in the context of frontal waves it is likely that the scale of the wave is important: features of the scale of cloud cells, $10km$ say, are likely to be 'triggered' phenomena (such as the rainbands of Locatelli *et al.* (1993)) but on synoptic scales moist adjustment is likely to be more pertinent.

Since frontal waves exist at the mesoscale, the question of how moist processes interact with a wave must take into account the lack of general understanding of the basic couplings and this should be borne in mind when generalising results. It would be insufficient to consider only one parametrization of the moist effects. At a further level, addressing the issue of forecasting of frontal waves, the position of the moist parametrizations used in operational NWP models within the theoretical heirarchy must be addressed.

Joly and Thorpe (1991) used a numerical normal mode type study on a time-evolving

basic state (so that the spatial structure obtained is constant but not necessarily exponentially growing), in which moist processes are allowed to occur in both the basic state front and the small amplitude perturbation. To summarise their findings for waves on moist Eady-wave basic states (such as those of Emanuel *et al.* 1987), short wave instability (along-front wavelength of around $1,000km$) was possible for slowly evolving but intense fronts, in which case mean e-folding times over a $24h$ period could be less than half a day. Once again, as shown by BTb for deformation frontogenesis and observed by ALJL in the ECMWF analyses, it seems that strong frontogenesis, in this case shearing deformation, is detrimental to frontal wave growth.

The existence of the short wave three dimensional modes was found by Joly and Thorpe (1991) to depend crucially on the presence of the diabatic heating. These modes were periodic in along-front direction (unfortunately the spatial structures were not shown) and it was commented that asymmetric structures, with narrower updraught regions, would be likely to experience faster growth. Study of such structures and the mechanism of interaction between diabatic heating and the wave would be valuable.

Much work on the effect of moist processes on cyclogenesis has focussed on 'primary' instabilities, for example the explosive synoptic scale systems observed to form off the East coast of the USA. Davis *et al.* (1993) studied model simulations of three extratropical cyclones; one over the continental USA and two over the West Atlantic. They found that although the pv anomalies attributable to the diabatic source may account for up to one third of the anomalous pv of the system (although there was much variation in the significance of latent heating between the cases), the fact that the anomalies are of relatively short horizontal scale means that they do not have a great vertical penetration, and hence have little influence on the baroclinic deepening. For secondary frontal cyclones this point applies regardless of the presence of moist processes, because it is partly their short horizontal scale which defines these phenomena. Davis *et al.* (1993) do speculate that one indirect effect of these pv anomalies would be to modify the vertical motion field in a feedback with the heating.

Since the modelled cyclones of Davis *et al.* (1993) showed reasonable agreement with surface rainfall observations, there is some confidence that the integrated diabatic sources were well represented in these studies. However, the authors do discuss how different convection schemes, in modifying the vertical distribution of heating and consequently the height of the pv anomalies,

can provide noticeably different degrees of surface deepening.

Reed *et al.* (1993b) show a difference due to convective parametrization scheme in a sensitivity study on simulations of the IOP5 storm from the ERICA project, a field study conducted off the East coast of the USA. The principal results of Reed *et al.* (1993b) are that the moist processes are crucial to the explosive deepening of the storm: the absence of moisture gives a storm which is slower (as follows from arguments of Parker and Thorpe (1995)) and shallower, by $19mb$, than the control simulation (a ratio of deepening of 2.58 between control and dry simulations). However, as found by Davis *et al.* (1993), the surface temperature features were not significantly modified by the absence of moisture. Removal of the surface fluxes from the simulation produced a system which was likewise slower and was shallower (by $6mb$) than the control; this is linked to the reduction in latent heating. Removal of the surface fluxes also *intensified* the surface thermal pattern: the fluxes act to warm the cold air and may cool the warm flow (some downward heat fluxes are seen in the control simulation).

Reed *et al.* (1993b) make an interesting comparison of various other studies of explosive cyclogenesis, calculating for each case the ratio of the control pressure drop and that computed without surface fluxes, $\frac{\Delta p(control)}{\Delta p(no fluxes)}$. This ratio can be negative or positive, for different storms, and Reed *et al.* (1993b) ascribe this to the presence or absence of negative surface heat fluxes in advance of the cyclone. In the ERICA IOP5 case, there are positive fluxes in advance of the system and these fluxes enhance the deepening (through subsequent latent heat release and a maintenance of the thermal advection). Neiman and Shapiro (1993a) attempted to quantify the surface fluxes in a rapidly deepening system off the East coast of the USA and found that the occurrence of negative fluxes to the East of the system coincided with the end of cyclogenesis. Again, the precise mechanism of this phenomenon is not explained (nor, at a more direct level, is the coupling with latent heating), but as the magnitude of the positive boundary-layer heating was diagnosed to reach orders of $5Kh^{-1}$ (albeit in a system where winter continental air passes over warm Gulf Stream water), the phenomenon would seem to be extremely important. In other cases where negative fluxes occur ahead of the cyclone, they act to weaken the system. An extension of this idea would be to explore how the surface fluxes bear on the behaviour of a warm conveyor belt in a mature frontal cyclone.

Balasubramanian and Yau (1994) used a 2-layer model to simulate explosive moist cyclo-

genesis on a baroclinic zone of around $3,000km$ extent (i.e. with a view to understanding explosive primary cyclogenesis), using a parametrization of slantwise convection. The model did not show great differences from a dry simulation until it reached a stage at which nonlinear advection produced a 'bent back warm front'. At this time, a surge of cold advection occurred at the mid-level (in the 2-layer model). This cold advection corresponded to a fall in the upper level height which produced an explosive deepening in the surface pressure. Balasubramanian and Yau (1994) note that it is the *frontogenesis* process which rapidly enhances the cold advection, giving an upper response: upper/lower level interactions are crucial to this model, as are interactions between the latent heating at the front and the cyclone. This is an example of the scale interactions induced by moist dynamics, as discussed by, for example, Joly and Thorpe (1989). Moist processes tend to reduce the horizontal scale of the fields of vertical motion and hence of the diabatically produced pv fields. Inversion of these pv fields, however, still tends to be dominated by the dry Rossby radius, so that narrow bands of convection can have a broad influence, as in Balasubramanian and Yau's (1994) model.

The above studies were performed with a view to describing moist influences on synoptic-scale cyclogenesis. Although much insight has been gained into these systems, there is still a lack of understanding of the mechanisms of rapid growth, particularly in the interaction with latent heating which, it appears, can have a variable degree of influence from case to case. Furthermore, work needs to be done if a clear picture of cyclogenesis on all smaller scales is to be obtained.

5 Outstanding issues

- Analysis of the IOP3 warm frontal rainband
- Origin and role of cold/dry intrusions
- Role of frontolytic, diffluent flow in nonlinear instability
- The effect of scale on the influence (and nature) of moist processes
- The degree of influence which moist processes can bear

- Influences of the boundary layer and surface fluxes on shallow instabilities

In order to address some of the above issues, it is proposed to perform numerical simulations of frontal waves using semigeostrophic and primitive equation models (such as that used by Malardel *et al.* 1993). The principal questions to be addressed by this work are:

1. What are the mechanisms by which moisture interacts with the dynamics of pv-strip instability?

2. What is the influence of horizontal scale of the wave on the instability? How does scale affect the interaction with moisture?

3. How do boundary processes (drag, heat fluxes, SST) modify the development?

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